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Decadal slowdown of a land-terminating sector of the Greenland Ice Sheet despite warming

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Ice flow along land-terminating margins of the Greenland Ice Sheet (GrIS) varies considerably in response to fluctuating surface meltwater inputs to the ice-sheet bed which lubricate the ice-bed interface, resulting in periods of faster ice motion[1,2]. Stronger melting results in faster ice motion during summer, but slower motion over the subsequent winter due to the evolution of a more extensive drainage system at the ice-sheet bed, which drains high-pressure regions more efficiently[2,3]. However, the impact of hydro-dynamic coupling on ice motion over decadal timescales remains poorly constrained. Here we show that annual ice motion across a 8000 km^2 land-terminating region of the west GrIS margin extending to 1100 m asl was 12% slower in 2007–2014 compared to 1985–1994, despite a corresponding 50% increase in surface meltwater production. Our findings suggest that hydro-dynamic coupling in this section of the ablation zone resulted in net ice motion slowdown over decadal timescales — not speedup as previously postulated[1]. Increases in meltwater production from projected climate warming may therefore further reduce the motion of land-terminating margins of the GrIS, which suggests that these sectors of the ice sheet are more resilient to the dynamic impacts of enhanced meltwater production than previously thought.

The Greenland Ice Sheet (GrIS) is losing mass at an accelerating rate[4,5] as a result both of increased surface melting[6] and enhanced ice discharge from accelerating marine-terminating glaciers[5]. Enhanced melting accounts for ∼60% of GrIS mass loss since 2000[5]; summer air temperatures over the south-west GrIS warmed by 0.9°C during 1994–2007[7] and meltwater production during the summers of 2007–2012 (except 2009) is without precedent in the last 50 years of reanalysis-forced reconstructions[8]. During 1993–2012 the average annual melt doubled from that which occurred during 1961–1990[8].

While the acceleration of marine-terminating glaciers is believed to be driven primarily by processes operating at the ice-ocean interface, atmospheric forcing can change ice motion at both land- and marine-terminating glaciers through the delivery of surface meltwater to the ice-sheet bed[1,9]. Surface meltwaters can drain rapidly to the ice-sheet bed via moulins and supraglacial lake drainage events which provide direct surface-to-bed connectivity and a mechanism by which surface meltwater can influence basal motion[10–12]. It was hypothesised that this mechanism could lead to a positive feedback between enhanced surface meltwater production and ice-sheet motion as ice would move more quickly to lower elevations where temperatures are warmer[1,13].

More recent studies have highlighted the importance of the subglacial drainage system in controlling the relationship between surface melting and ice motion through changes in system capacity and morphology[14–16]. During summer, rapid increases in meltwater from the ice-sheet surface result in periods when the subglacial drainage system is more highly pressurised, leading to transient periods when water pressure exceeds ice overburden pressure, resulting in enhanced basal sliding[15]. However, subglacial drainage system capacity increases in response[17,14], introducing a negative feedback which lowers the water pressure and reduces basal sliding[16,18]. By the end of summer, an efficient drainage system has evolved upglacier[15,16] which drains surrounding regions of the ice-sheet bed that were previously hydraulically isolated. This reduces basal lubrication during the subsequent winter, counteracting summer speed-up and making net annual ice motion relatively insensitive to summer melting[2,3].

Despite these advances in understanding coupled hydro-dynamics, it remains unclear whether enhanced surface melting has a long-term impact on annual ice motion. Eight Global Positioning System (GPS) stations on a transect extending 130 km inland in the south-west GrIS showed an average 10% decrease from 1991 to 2007, during a period when surface melt increased significantly, but with considerable spatial variability[19], and the slowdown trend at lower sites continued into 2012[20]. Meanwhile from 2009 to 2012 a small acceleration signal was observed above the ∼1500 m asl equilibrium line altitude[21]. The parametrization of basal lubrication in higher-order ice-sheet models, using observations from south-west Greenland, suggests that basal lubrication is unlikely to increase the contribution of the ice sheet to sea level rise by more than 5% of the contribution expected from a negative surface mass budget alone, and could conceivably act as a negative feedback upon ice motion[22].

Here we present observations of annual motion spanning three decades, which extend back to 1985. Our ∼8000 km^2 study area extends along ∼170 km of predominantly land-terminating margin of the west GrIS to ∼50 km inland and ∼1100 m asl (Fig. 1). We apply feature tracking (see Methods) to 475 pairs of remotely-sensed optical Landsat imagery separated by approximately one year[23]. Next, we derive robust ice motion and uncertainty estimates over ∼1–2 year periods from 1985 to 2014 (Fig. 2b and Extended Data Fig. 4), and over multi-year reference periods spanning (a) 1985–1994, capturing the period before air temperatures began to warm[7], and (b) 2007–2014, corresponding to the recent series of record melt summers[8].

Ice motion shows a clear regional slow down (Fig. 1) with 84% of the study area flowing more slowly in 2007–2014 than in 1985–1994 (Fig. 1a). On average ice motion slowed by 12% across the study area. Slowdown was strongest (∼15-20%) at elevations below ∼800 m asl (Fig. 1b). Isolated areas experienced speedup in 2007–2014 compared to 1985–1994. In the far north-east, the speedup can likely be attributed to the dynamics of the neighbouring marine-terminating Jakobshavn Isbrae, which like many of Greenland's marine-terminating glaciers has accelerated since the mid 1990s[5].

The ice motion record can be divided into two statistically significant periods (see

Methods). Segmented linear regression ($R^2 = 0.79$) shows that there was no significant trend in ice motion during 1985–2002 ($p = 0.85$). Ice motion slowdown likely began around 2002 and despite inter-annual variability, there was a robust overall trend of -1.5 m yr^{-2} during 2002–2014 ($p < 0.01$). Meanwhile, surface meltwater production (Fig. 2a) can be divided into three statistically significant periods (see Methods): sustained 'low' melt of 2.1 water equivalent (w.e.) m yr^{-1} during 1985–1993, rising melt during 1993–2002, and sustained high melt of 3.2 w.e. m yr^{-1} during 2002–2014 coincident with when ice motion began to slow down. Overall there was a 49.8% rise in surface meltwater production across our study area between 1985–1994 and 2007–2014.

We explored temporal variability in ice motion along three transects (Fig. 1) selected to represent different ice-marginal conditions. Transect A (Fig. 3a) extends 80 km inland from Nordenskjöld glacier, which has a lacustrine-terminating margin; transect B (Fig. 3b) extends ~30 km inland from a land-terminating margin; and transect C (Fig. 3c) extends ~30 km inland from the marine-terminating Alangordliup sermia. Transects A and B slowed down during 2000–2014 to attain velocities averaged along the transect 19% and 18% lower in 2013/14 than during the 1985–1994 reference period respectively. Ice motion characteristics at the marine-terminating transect C were more complex. The transect slowed on average from the mid 2000s to 2014 although ice motion within 10 km of the margin sped up in the late 2000s following earlier slowdown and by 2013–2014 was flowing up to ~ 50 m yr^{-1} faster than during the 1985–1994 reference period. Such behaviour is in line with other tidewater glaciers that have recently accelerated[5].

The slowdown signal across our predominantly land-terminating region extends up to ~1100 m asl (Fig. 1) where the mean ice thickness is ~850 m[24]. The clear deceleration in ice motion requires a decrease in rates of either internal ice deformation, basal motion or a combination of both mechanisms. Melting has caused marginal thinning of the GrIS[25–27]. During 1993–1998 land-terminating glaciers on the west GrIS margin thinned by 0.02-0.23 m yr^{-1} below 1000 m asl[25]. Our study area thinned by $\sim$0.2 m yr^{-1} during 2003–2007[26] and this rate increased to 1–1.5 m yr^{-1} during 2011–2014[27]. We modelled the velocity change that would be caused by 10–20 m of ice thinning (and the associated gradient changes) along transect A over the 1985–2014 study period (see Methods), corresponding to a maximum thinning rate of $\sim$0.6 m yr^{-1}. The resulting change in driving stress can explain only $\sim$17-33% of the observed overall 12% slowdown signal beyond 10 km from the ice sheet margin and can explain none of the slowdown beyond 50 km from the margin (Extended Data Fig. 5c). Thus, while a component of the observed slowdown can be explained by changes in driving stress through ice thinning, the majority of the slowdown (i.e. the remaining 67–83%) must be the result of processes operating at the ice-bed interface causing a reduction in basal motion.

Previous studies have suggested that the coupling between surface melting and basal motion is self-regulating, such that there is no significant relationship between melting and ice motion over annual timescales[2,20]. In agreement with these studies we find no relationship between annual melt volume and annual ice motion ($R^2 = 0.08$). There is however a strong relationship between antecedent melt volumes and ice motion (Extended Data Table 1). The mean melt volume from each observation period and previous year combined explain 23% of observed ice motion ($p < 0.05$), increasing to 44% when the previous four years of melt are included. Moreover, melt volumes explain 50% of observed ice motion when the mean melt volume is calculated using only the previous three years ($p < 0.01$).

We therefore hypothesise that sustained high surface meltwater production (Fig. 2a) is responsible for the observed slowdown. Observations from the GrIS show that during the melt season, the large cumulative increase in the rate of meltwater supply to the ice-sheet bed results in the expansion of a channelized subglacial drainage system[16], even beneath ice ~1 km thick[11]. As air temperatures warm, more meltwater at higher elevations allows an efficient drainage system to evolve which extends further into the ice sheet. Summers of extreme melt result in a higher-capacity, more extensive channelized drainage system which stays open at atmospheric pressure for longer after cessation of melt[2,3]. Dye tracing of alpine glaciers[28] indicates that transit speeds through unchannelized drainage systems are ~0.01 m s^{-1} (~850 m d^{-1}). Thus, while channels at atmospheric pressure beneath ~1 km thick ice close within hours to days[11], channels which stay open, for example for just two days as opposed to one, have the capacity to evacuate significantly more water from surrounding unchannelized regions of the ice-sheet bed, causing more widespread dewatering of the ice-bed interface. Sources of meltwater, including frictional melting by basal slip and geothermal heat, will enable water pressure (P_w) to recover gradually through the subsequent winter, but may be insufficient to replace the stored waters evacuated during the previous melt season.

Previous observations have illustrated the importance of changing connectivity between channelized and unchannelized regions of the ice-sheet bed in controlling ice velocities late in the melt season[12]. We postulate that these unchannelized drainage regions and their connectivity to the channelized drainage system govern ice motion not only late in the melt season but also during the following winter and spring. We hypothesise that if increases in drainage efficiency occur year-on-year, gradual net drainage of water stored in unchannelized regions of the ice-sheet bed will result in reduced basal lubrication and net ice slowdown. Additional field observations, such as borehole arrays transverse to subglacial channels recording water pressure gradients (e.g.[29]), in conjunction with hydrological modelling (e.g.[14]) are required to test the robustness of our hypothesis. Furthermore, while melt-driven seasonal evolution in subglacial drainage can impact the flow of tidewater glaciers[9], their ongoing acceleration[5] during a period of warming, in contrast to our observations, suggests that other processes are controlling their dynamics.

Our observations of GrIS ice motion made over three decades provide conclusive evidence that a 50% rise in meltwater production has not led to ice speedup along a land-terminating margin; instead average annual ice motion at elevations below 800 m asl slowed by >15% and likely by at least 5% up to 1100 m asl. Only $\sim$17-33% of the slowdown can be explained by reduced internal deformation caused by ice thinning, and we therefore hypothesise that since 2002, increases in subglacial drainage efficiency associated with sustained larger melt volumes have reduced basal lubrication, resulting in slower ice flow. It remains unclear whether the observed slowdown occurs at elevations above 1100 m asl and whether the slowdown will migrate inland as enhanced melting extends to higher elevations and allows a more extensive efficient subglacial drainage system to evolve. Furthermore, while our findings relate to land-terminating margins, the forcing mechanisms which have driven the recent speedup of many tidewater glaciers remain poorly understood[5,26] and require a similar examination of annual ice motion over decadal timescales.

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Author Contributions: A.J.T., P.W.N. and N.G. designed this study. A.D., N.G. and A.J.T. developed the processing chain used for feature tracking of Landsat imagery.

A.J.T., A.D. and N.G. processed the Landsat imagery. A.J.T. and D.G. calculated the impact of changing ice geometry upon ice motion. E.H. processed the melt data. A.J.T., N.G. and P.W.N. analyzed the results. A.J.T., P.W.N. and N.G. wrote the manuscript. All authors discussed the results and edited the manuscript.

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Figure 1: Study area in the ablation zone of the western GrIS. Colour shows percentage change of ice velocities in 2007–2014 reference period compared to 1985–1994 reference period (see main text). Transects correspond to Fig. 3. Ice surface contours from Howat *et al.*[30]. Gray denotes areas where ice velocities cannot be resolved; green denotes land areas; light blue denotes inland and coastal waters. Inset (a): percentage change of ice velocities in 4% bins. Inset (b): Median percentage change in each 100 m elevation band between 400–1100 m asl and associated uncertainties (see Methods).

Figure 2: Surface melting and ice motion averaged over the study area. (a) Annual mean modelled surface melt (gray), smoothed with a 5-year moving mean (black), both in water equivalent (w.e.) m per year (see Methods). (b) Median ice velocities during each period (black boxes) calculated using the common sampling pixels across the time series, $\pm 1\sigma_p$ (see Methods). The width of each box corresponds to the total timespan of the pairs of Landsat images acquired during each period. The height of each box corresponds to $\pm \sigma_p$ (see Methods). Blue and red lines illustrate the trends in ice velocity computed by segmented linear regression weighted by σ_p. (c) The altitudinal distribution of the common sampling pixels used to compute the velocities in (b).

Figure 3: Ice velocities along three transects in the study area. The transects correspond to those shown in Fig. 1. Only periods in which ice velocities are observed along at least 60% of each transect are shown. Velocities during the 1985-1994 reference period are also shown. (a) Transect A, (b) Transect B, (c) Transect C.

Methods

Remote sensing of ice motion. We applied feature tracking techniques to extract ice motion from Landsat Program imagery. Landsat images were obtained from the U.S. Geological Survey (via the earthexplorer.usgs.gov catalogue) and the European Space Agency (via the EOLI-SA catalogue). Here we provide an outline of the processing strategy and the specific parameters used in this study. A detailed description of the processing strategy is available elsewhere[23]. Our approach builds individual annual velocity fields from feature tracking of Landsat pairs, overlapping in time and space. These velocity fields are then combined over inter-annual time periods in order to increase the robustness of the velocity estimates and to enable statistical determination of uncertainties.

We used images from the Landsat 5, 7 and 8 missions; the quality and quantity of images from Landsat missions 1–3 were insufficient to permit their inclusion in this study. We identified 475 image pairs with temporal baselines between 352–400 days acquired during April–October over the 1985–2014 study period (see Supplementary Information). The temporal baseline of $\sim$1 year was chosen to minimize the impact of seasonal flow variability upon inter-annual trends in ice velocity, as we are specifically interested in long-term changes in ice motion.

To enhance the images prior to feature tracking we used Principal Component Analysis to combine the optimum spectral bands, identified during testing as bands 2 and 3 for each satellite mission. A high pass filter (utilising Sobel kernels) was used to compute the intensity gradients of each image, enhancing surface features such as crevasses and reducing the impact of basal topography related features which by definition are temporally stable. We then used the gradients as features to be tracked. The tracking (used to extract ice displacement) was performed on matching windows of 44 pixels (1320 m) and a grid spacing of 8 pixels (240 m), whilst the search window was set automatically to correspond to the maximum expected displacement over the baseline duration between the two images based on previous velocity observations[31].

The processing strategy exploits the redundancy offered by multiple, spatio-temporally overlapping pairs to efficiently remove velocity outliers and produce robust velocity fields. Firstly, we filter low quality velocity estimates by applying a threshold to the signal-to-noise ratio returned during the feature tracking. The threshold value was determined by examining all velocity pairs to identify the value beyond which the Median Absolute Deviation (MAD) of stable area velocities became asymptotic. We use a median based approach to minimize the impact of outliers and because the distribution of velocity tends not to follow a normal distribution[23]. Next, for the period 2000–2014, velocities were grouped into 1-year time periods, while for the period 1985–2000, velocities were grouped into 2-year time periods due to the lower number of Landsat pairs available. This provides spatio-temporal redundancy in the velocity estimates at each pixel and enables us to quantify uncertainties. To produce the final velocity field for each period we compute the median of all the available velocity estimates at each pixel. Lists of the Landsat pairs which contribute to each period are available in the Supplementary Information. Finally we compute the 1σ uncertainty of the velocity estimate at each pixel in each period. To do so we fit a law of the form

$$\sigma = \frac{k}{2} \frac{MAD}{N^{\alpha}}, \tag{1}$$

where N is the number of velocities used to compute the median velocity, σ is the 1-σ confidence interval and k and α are the parameters to be determined. k and α

were determined for each time period from the velocity estimates made over stable land areas whose true value is known and equal to 0. Uncertainty of the final ice velocity of each pixel at each time period is then obtained by extrapolating this relationship to on-ice areas with the appropriate values of MAD and N for a given pixel. Considering N and MAD at the pixel level allows the surface characteristics at the location of the pixel considered (e.g. surface conditions, variability of velocity during the time period considered) to be taken into account. Pixels with σ >60 m yr^{-1} are discarded in the subsequent analysis.

We computed the percentage change in ice velocities between 1985–1994 and 2007–2014 at all of the pixels which are common to both periods (Fig 1) and then computed the median percentage change, both over the whole study area and in 100 m asl elevation bands. For each elevation band in Fig. 1b we calculated the uncertainty of the percentage change by firstly estimating the uncertainty of the 1985–1994 (e_1) and 2007–2014 (e_2) velocities separately as $\sqrt{\sum_{i=1}^{N} \sigma_i^2/N}$, and then computing $\sqrt{{e_1}^2 + {e_2}^2}$. Residual striping patterns in Fig. 1 are caused by lines of missing data in the Landsat 5 imagery.

To compute inter-annual median velocities (Fig. 2b), firstly periods in which less than 30% of the study area have observations were discarded. Then, the pixels common to all the retained periods (shown in Extended Data Fig. 4) were selected so as to avoid temporal variation caused by spatial bias. For each period we calculated the median of the 8025 temporally common pixels and the associated uncertainty, estimated as $\sigma_p = \sqrt{\sum_{i=1}^{N} \sigma_i^2/N}$. The altitudinal distribution of the common sampling pixels is shown in Fig 2c.

The decision to discard periods in which less than 30% of the study area has observations is a compromise between calculating the median velocity of each period using the greatest possible number of pixels common to all periods, versus retaining the maximum possible temporal resolution. We examined the sensitivity of the ice motion time series (Fig. 2b) to 40% and 50% thresholds. At 40%, there are 5970 more common sampling pixels than at 30%, but temporal resolution decreases as ice velocities observed in 2003–2006 are no longer retained. The R^2 of the two-trend model decreases to 0.65. The rate of slowdown during 2002–2014 increases from 1.5 m yr^{-2} to 1.9 m yr^{-2} ($p < 0.01$). At 50%, there are 8397 more common sampling pixels than at 30%. Velocities observed during the 1995–1997 period are also discarded. The R^2 of the two-trend model is 0.63. The rate of slowdown during 2002–2014 remains the same as at 40% ($p < 0.01$). There is no statistically significant trend in ice motion from 1985 to 2002 ($p > 0.05$) at either of the tested thresholds. From this sensitivity analysis we conclude that our 30% threshold case yields the highest temporal resolution and also the most conservative trend in ice motion during 2002–2014.

Identification of trends in melting and ice motion. For both time series we test whether they can be divided into temporally distinct populations separated by break dates. We therefore apply the Mann-Whitney-Wilcoxon (MWW) rank sum test in order to investigate the effect of prescribing different break dates. We chose the MWW test over the t-test as we do not know whether the dataset follows a normal distribution. The test computes the probability that the populations, separated by prescribed break dates, are similar.

The melt time series (Fig. 2a) consists of several years of sustained low melt, followed by a period of rising melt and then several years of sustained higher melt. We therefore test whether the melt time series can be split into three statistically different populations, with two break dates separating the periods of sustained low, rising and sustained high melt. We find that the break date combinations of (i) 1992 and 2001, (ii) 1993 and 2001, and (ii) 1993 and 2002 are all significant with 95% confidence

(Extended Data Fig.2). To find the best possible combination of break dates, we then compute the root-mean-squared error, or residuals, of the best fitting 3-trend segmented linear regression model (Extended Data Fig. 2). We observe the lowest residuals for break dates of 1991–1993 and 2001–2004. The combination of break dates which satisfy the MWW test and have the lowest residuals is 1993 and 2002. With this chosen combination of break dates, the probability that (i) 1985–1993 is similar to 2002–2013 is 0.02%, (ii) 1985–1993 is similar to 1994–2001 is 2%, and (iii) 1994–2001 is similar to 2002–2013 is 3%.

For the ice motion time series, we first test whether it can be divided into two temporally distinct populations, in order to justify the use of segmented linear regression (Fig. 2). We apply the MWW test as previously. We find that for break dates beyond mid 2001, the null hypothesis (equal median) can be rejected (Extended Data Fig. 3a), meaning that the pre- and post- 2001 populations are statistically different with 95% confidence. To test the ability of two distinct periods separated by a given break date to represent the velocity time series we then compute the residuals of the best fitting 2-trend segmented linear regression model for a set of break dates spanning the time period of the dataset (Extended Data Fig. 3b). We observe a minimum for a break date in 2002 but with a region of low residuals spanning the period 1998 to 2004. We conclude from these tests that there are two distinct temporal populations of ice motion in our dataset and that the break point occurs during 1998–2004. For our analysis (Fig. 2) we select a break date of 2002, which corresponds both with the lowest residuals and the MWW test suggesting that the pre- and post-2002 populations are statistically different.

We then examine whether there is a statistically significant relationship between meltwater production and ice motion by applying linear regression analysis to investigate the extent to which variability in ice motion can be explained by (a) temporally coincident and (b) antecedent meltwater production (Extended Data Table 1). In (b) we quantify antecedent meltwater production in two different ways. In one scenario we calculate the mean during the period of observed ice motion and the preceding N years. In the other scenario we calculate the mean of only the preceding N years.

Impact of varying baseline durations on annual velocity. Images separated by baseline durations of 352–400 days duration were paired together for feature tracking. Here we examine the impact that the variable baseline duration has on the velocity field of each period.

The average start day-of-year of the pairs which comprise each period are shown in Extended Data Fig. 1a. The start day becomes less variable once imagery from Landsat 7 comes online in 1999. The average baseline duration increases by $\sim$15 days after 1999 (Extended Data Fig. 1b), increasing the proportion of the baseline attributable to summer motion (defined as 1 May to 31 August in common with previous studies[2,3]) by $\sim 2\%$ (Extended Data Fig. 1c).

We use mean summer and winter velocities from Leverett Glacier sites S1 to S4 during 2009 to 2012[2,3] to test the sensitivity of annual velocities to the varying baseline duration. Winter days in each period are ascribed velocities of 81.6 m yr^{-1} and summer days are ascribed velocities of 127.6 m yr^{-1}. We then estimate the mean annual velocity that would be expected for each period (Extended Data Fig. 1d). Variations in the baseline duration between periods are estimated to impact extracted annual ice motion by <2 m yr^{-1}. Furthermore, according to this analysis, increased baseline durations during the 2000s leads to an small artificial increase in ice motion caused by the feature tracking method. This is in the opposite direction to the inter-annual slowdown signal which we observe in our study area and leads us to conclude that our slowdown trend is robust to varying baseline durations.

Impact of changing ice geometry on velocity. During the last 30 years, the Greenland ice sheet (GrIS) has thinned along its margins[32,26,27], changing the geometry of the ice mass. It is likely that these changes will have affected ice velocity by modifying driving stress. Here we evaluate the impact that thinning may have had on velocity along transect A (see Fig. 1), in order to bound the extent to which our observed slowdown could be the result of geometric changes.

We characterize the surface velocity u_s as a sum of a basal sliding (u_b) and vertical shear deformation (u_d) contributions[33]:

$$\begin{aligned} u_s =& u_b + u_d \\ =& C_b(\rho_i g S H)^m + \frac{A}{4}(\rho_i g S)^3 H^4, \end{aligned} \tag{2}$$

where A is a temperature-dependent Glen's flow parameter, ρ_i is the density of ice, g is gravitational acceleration, S is the surface slope (positive where the surface lowers toward the margin), H is the ice thickness, and C_b and m are parameters related to basal sliding. A, C_b and m are in general poorly-constrained, and are likely to vary spatially; however, the only assumptions we make in our analysis regarding these parameters are that they do not change significantly over the time interval of interest, and further that m is less than or equal to 3. Note that our model allows for either the power-law rheological model of Weertman (1957)[34] or the Newtonian till model of Alley (1987)[35]. Thus the maximum deceleration predicted by the above model bounds the slowdown that can be explained by geometric changes alone. Below, we estimate this maximum deceleration to first order.

We introduce the variable λ, which represents the fraction of surface velocity explained by vertical shear, i.e.

$$u_d = \lambda u_s, \quad u_b = (1-\lambda) u_s.$$

If we consider a small change δS in slope ($\delta S \ll S$), and a small change δH in ice thickness ($\delta H \ll H$), Eqn. 2 leads to the following change in u_s:

$$\begin{aligned}\frac{\delta u_s}{u_s} =& \left(\frac{m u_b + 3u_d}{u_s}\right)\frac{\delta S}{S} + \left(\frac{m u_b + 4u_d}{u_s}\right)\frac{\delta H}{H} + \mathcal{O}(\delta H^2, \delta S^2) \\ =& (m(1-\lambda) + 3\lambda)\frac{\delta S}{S} + (m(1-\lambda) + 4\lambda)\frac{\delta H}{H} + \mathcal{O}(\delta H^2, \delta S^2),\end{aligned} \tag{3}$$

where the $\mathcal{O}$-notation is employed to signify terms which are of order δH^2 and δS^2 or higher and thus negligibly small. Again, we make no assumption regarding the spatial variability of λ other than the fact that it is between 0 and 1, as our aim is to find the conditions under which velocity is most sensitive to thinning. With $\left(\frac{\delta S}{S}\right)$ positive and $\left(\frac{\delta H}{H}\right)$ negative in Eqn. 3, then at any point along the transect, and for any $m \leq 3$, the change to u_s cannot be more negative than when the flow is due to vertical shear, i.e. when λ is equal to 1. The first-order relative change in surface velocity, $\frac{\delta u_s}{u_s}$, is thus bounded by

$$3\frac{\delta S}{S} + 4\frac{\delta H}{H}. \tag{4}$$

We use Expression 4 to estimate the maximal impact of these thinning scenarios on ice velocity, solving every 240 m along transect A. To estimate the total ice thinning during 1985–2014, $\delta H = 10, 20$ are prescribed at the ice sheet margin and linearly interpolated along the transect to δH = 0 m at 100 km inland (equivalent to the equilibrium line altitude, $\sim$1500 m asl[36]) (Extended Data Fig. 5a). We add δH to current ice thickness along the transect[24] to set H to values appropriate for 1985. We prescribe the initial slope S as the mean slope in our study area, 0.02 m/m. The change in slope, δS, is calculated from the prescribed linear change in ice thickness over distance inland.

In Extended Data Fig. 5b we plot Expression (4) corresponding to the two thinning scenarios described above. These profiles represent the largest (i.e. most negative) percentage changes in velocity associated with the prescribed geometric change. We then convert these profiles to the predicted reductions in velocity and remove them from the observed 1985–1994 velocities (Extended Data Fig. 5c), generating a lower bound for the 2007–2014 velocities under the assumption that the observed slowdown was geometrically induced. Between $\sim$0-5 km from the ice margin for the $\delta H = 20$m scenario, the observed slowdown is within the range predicted by Expression 4, and we cannot reject the possibility that the thinning here was responsible for the slowdown. However, between 10–50 km inland, at most 17–33% of the observed slowdown can be attributed to changing ice sheet geometry depending on the prescribed δH. By 60 km inland, there is essentially no net change in ice velocity attributable to geometrical changes.

It remains to consider our assumption that the Glen's law parameter A can be treated as constant in time. Phillips et al[37] demonstrated that latent heat transferred to the ice from surface melt could warm glacial ice at depth, thereby leading to an increase in A. However, this process would result in acceleration rather than deceleration. Thus our decision not to consider temporal changes in A in our analysis of maximal geometrically-induced slowdown is justified. We conclude that the slowdown which we have observed during 1985–2014 is not explicable by geometrical changes to the ice sheet alone and instead must be dominated, at distances greater than $\sim$5 km from the margin, by other processes impacting basal motion.

Surface melting. GrIS annual melting was output from a runoff/retention model applied to downscaled ERA-I data on an equal-area 5x5 km polar stereographic grid for the Greenland region[38]. We calculated mean annual melt rates for the study area (67.45°N, 51.5°W to 69.2°N, 49.2°W). Inter-annual fluctuations and trends from several independent melt models show good agreement[39], including with the methodology used in this study.

Code availability. The feature tracking algorithm is proprietary software developed and licensed by GAMMA Remote Sensing. Requests for the code underlying the processing strategy used in this study should be sent to its original authors[23]. Similarly, requests for the code constituting the runoff/retention model should also be addressed to its original authors[40].

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Extended Data Figure 1: Sensitivity of extracted ice motion to variations in baseline duration. For each period (a) the average start Day-Of-Year of all pairs used in the period; (b) the average baseline duration of all pairs used in the period; (c) the proportion of the baseline duration attributable to summer, which is defined as 1 May to 31 August; and (d) the annual velocity which would be expected in the ablation zone of the Leverett glacier catchment based on the average proportion of summer versus winter and the average baseline duration for each year.

Extended Data Figure 2: Statistical significance of three different periods of surface meltwater production. Hypothesis test of the Wilcoxon rank sum test at 95% confidence, showing that three periods of surface melt separated by the specified dates have statistically different medians (outlined white region). The shading shows the residuals of a 3-trend linear segmented regression model fitted to melting at each possible combination of break dates, expressed as the root-mean-square error.

Extended Data Figure 3: Statistical significance of two different periods of ice motion. (a) Hypothesis test of the Wilcoxon rank sum test for equal medians, testing the probability of the two populations separated by the specified date to be similar, with 95% confidence. 0 signifies that the hypothesis of equal medians cannot be rejected and 1 signifies that the hypothesis of equal medians can be rejected. (c) Residuals shown as the sum-of-squares $(\text{m yr}^{-1})^2$ of a 2-trend model fitted to velocities at each possible break date.

Extended Data Figure 4: Ice velocities during each period. The velocities have uncertainties < 60 m yr^{-1} and were observed across at least 30% of the study area in each period (see Methods).

Extended Data Figure 5: Impact of changing ice geometry on ice motion. Ice thinning of 10 m (purple) and 20 m (blue) at the ice margin through to 0 m at 100 km inland were applied to Transect A. (a) Prescribed change in ice thickness over transect length. (b) The ratio of velocity change calculated by Eqn. 4. (c) Left axis: Observed ice velocity during 1985–1994 (dotted green) and 2007–2014 (dotted red). Modelled velocities in 2014 (solid lines) for the prescribed ice thicknesses. Right axis: ice thickness (dashed gray)[24].

Extended Data Table 1: Statistical relationship between melting and ice motion. The results of linear regression analysis carried out between all periods of ice motion in Fig 2b and different estimates of temporally coincident[(1)] and antecedent meltwater production (see Methods).